P3.7 TROPICAL CONVECTION'S ROLES IN TROPICAL TROPOPAUSE CIRRUS

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1. INTRODUCTION

Remote sensing observations reveal the frequent occurrence of tropopause cirrus, thin cirrus layers located near the tropical cold-point tropopause at altitudes between 15 and 18 km. These layers typically are several hundred meters to one kilometer in thickness, extend several hundred to more than one thousand kilometers horizontally, and persist for time periods of several hours to several days. In addition, they are generally subvisible, or too thin to be detected visually from the surface.

The radiative heating associated with tropopause cirrus has important impacts on the thermal structure and vertical velocity in the upper tropical troposphere (McFarquhar et al., 2000). In addition it has been suggested that precipitation from tropopause cirrus plays an important role in producing the low water vapor mixing ratios observed in the lower stratosphere (Tsuda et al., 1994; Potter and Holton, 1995; Jensen et al., 1996a). Despite their importance to global climate, tropopause cirrus processes are poorly understood, in part because of a shortage of detailed observations of the properties of these clouds. *In situ* observation requires specialized aircraft that can operate near the tropical cold-point tropopause at altitudes of 16 to 18 km and in occasionally turbulent conditions. Such observations are available for only a very small number of cases (Heymsfield, 1986; Booker and Stickel, 1982).

Two primary hypotheses have been proposed for the formation of tropical tropopause cirrus layers. The first is that these layers are the remains of outflow anvils from deep convection. While this mechanism likely plays a role in some cases, several factors suggest that this in not the primary formation mechanism for tropopause cirrus. First, tropopause cirrus layers are frequently observed in regions far from organized deep convection. Second, recent work suggests that tropical convection is generally capped at altitudes 2-4 km below the tropical cold-point tropopause where these layers are found, with only very infrequent penetration to this level (Gettelman et al., 2001; Folkins et al., 1999; Keith, 2000; Highwood and Hoskins, 1998). Finally, results from a study of cirrus anvil decay conducted using a two-dimensional cloud resolving cirrus model suggest that, in the absence of large-scale forcing, thin cirrus layers that form from anvils are not able to survive for the long time periods that are observed (Boehm, 1999).

The second hypothesis for the formation of tropopause cirrus is that these layers form *in situ* as a result of slow, synoptic-scale uplift and cooling of a moist layer near the cold-point tropopause. This hypothesis is the focus of this abstract.

Three primary processes are involved in the *in situ* formation of tropopause cirrus: 1) moisture transport from the surface into the upper troposphere by deep convection, 2) additional vertical moisture transport to the cold-point tropopause by large-scale ascent, and 3) large-scale cooling and cirrus formation. In this abstract we present a conceptual framework in which tropical convection plays an important role in each of these processes. Tropical convection's role in the first process is obvious, and will not be discussed further. The second process is necessary since, as mentioned earlier, tropical convection is generally capped several kilometers below the cold-point tropopause.

Mechanisms by which tropical convection plays roles in processes 2 and 3 are presented in sections 2 and 3, respectively. The results are summarized in section 4.

2. VERTICAL MOISTURE TRANSPORT

Since recent work suggests that tropical convection and its associated moisture transport are generally capped 2-4 km below the cold-point tropopause, additional vertical moisture transport is required to explain the frequent presence of cirrus layers near the cold-point tropopause. A likely mechanism for accomplishing this transport is gentle, large-scale ascent in the tropical upper troposphere. Such large-scale ascent has been indicated by a wide variety of indirect measurements and estimates (Rosenlof, 1995; Eluszkiewicz et al., 1996; Rosenlof and Holton, 1993; Seol and Yamazaki, 1999; Mote et al., 1996, 1998; Niwano and Shiotani, 2001, Hall and Waugh, 1997; Boering et al., 1996).

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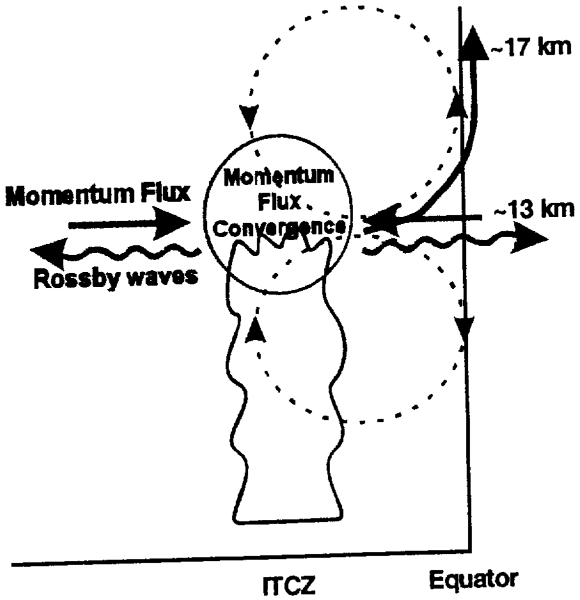

Figure 1: Schematic showing the generation of rising motion in the equatorial upper troposphere in response to momentum transport (green arrow) by Rossby waves (wavy red arrows) generated by tropical convection. The blue circle denotes the maximum in momentum flux convergence (see Figure 2) located in the tropical upper troposphere. The dashed lines indicate the meridional circulations that develop in response to this convergence to maintain thermal wind balance.

Despite important differences in the methodologies employed in these studies, the estimated vertical velocities in the upper tropical troposphere are very consistent, with values generally between 0.2 mm s^{-1} and 0.4 mm s^{-1}. This rising motion represents the lower extent of the Brewer-Dobson circulation, the large-scale circulation primarily confined to the lower stratosphere (Brewer, 1949; Dobson, 1956).

The processes responsible for this tropical upwelling are not well understood. It is partially driven by extratropical "wave drag" in the extratropical stratosphere and mesosphere. However, numerical experiments by Plumb and Eluszkiewicz (1999) suggest that extratropical wave forcing alone is unable to produce tropical upwelling of the observed magnitude. Therefore, other mechanisms must also contribute to the generation of rising motion in the tropical upper troposphere, and it s highly likely that these same mechanisms also transport moisture to the cold-point tropopause.

One possible such mechanism is closely connected with tropical convection (Boehm and Lee, 2002). In this mechanism, shown schematically in Figure 1, rising motion is generated in response to the meridional convergence of zonal Rossby wave momentum (eddy momentum flux convergence, hereafter) in the tropical upper troposphere. Tropical Rossby waves can be readily excited by the zonally asymmetric component of the diabatic heating associated with tropical convection (Gill, 1980). Under favorable conditions, these waves are able to propagate poleward into the extratropics (Hoskins and Karoly, 1981; Sardeshmukh and Hoskins, 1988). As the waves propagate away from their source region, wave breaking, nonlinearity, and other dissipation mechanisms including friction dampen the wave flux. The end result is that Rossby waves that are generated in the tropics pump westerly momentum into the tropics, leading to eddy momentum flux convergence in the tropical upper troposphere.

Analyzing 16 years of NCEP/NCAR reanalysis data, Lee (1999) showed that the net eddy momentum flux convergence is positive in the equatorial upper troposphere, indicating that in this region the wave momentum fluxes of tropical origin dominate those of midlatitude origin. Furthermore, it was shown that the spatial and temporal scales of the waves responsible for the momentum flux convergence are consistent with those of the El Niño-Southern Oscillation (ENSO) and the Madden-Julian Oscilation (MJO), suggesting that eddy momentum flux convergence in the tropical upper troposphere is ultimately driven by heating associated with tropical convection.

Figure 2 shows the mean January eddy momentum flux convergence field calculated based on NCEP/NCAR reanalysis for the period 1958-1997. Positive values indicate momentum flux convergence (eastward zonal wind acceleration), while negative values indicate momentum flux divergence (westward zonal wind acceleration). There are three main regions of momentum flux convergence: one in the mid-latitudes of each hemisphere and one near the equator. Momentum flux convergence in the mid-latitudes is primarily due to momentum transport by extratropical baroclinic disturbances, while momentum flux

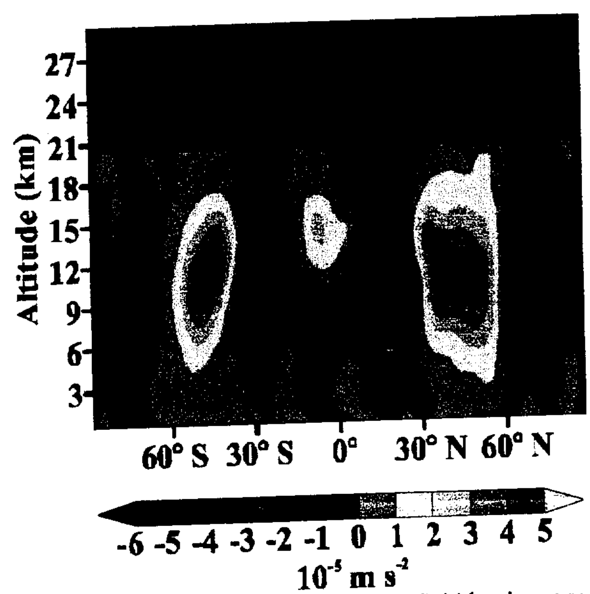

Figure 2: Momentum flux convergence field for January. See text for discussion.

convergence in the tropics is due to momentum transport by Rossby waves generated by tropical convection. There are four main regions of momentum flux divergence: one in the subtropics in each hemisphere and one at polar latitudes in each hemisphere. These are regions in which Rossby waves tend to break. The feature in this figure most relevant to this work is the tropical region of momentum flux convergence centered at a latitude of 8° S and at an altitude of 14 km. The location of this region is consistent with the mean latitude of the ITCZ during January and with observations that deep convection is generally capped at an altitude of 14 km.

This momentum flux convergence in the tropical upper troposphere produces a vertically varying eastward acceleration of the zonal wind that tends to destroy thermal wind balance in this region. In order to maintain thermal wind balance against this tendency, meridionally overturning circulations must develop. These circulations act to produce meridional gradients in temperature that are consistent with the differential zonal wind acceleration, and also to maintain the zonal wind at the steady state that is observed. Of particular interest is the rising motion that is predicted above the level of maximum momentum convergence near the equator, precisely where ascent is required to explain the tropical upwelling of the Brewer-Dobson circulation and the transport of moisture to the cold-point tropopause.

To test this hypothesis, model runs were conducted using the axisymmetric version of a multi-level primitive equation model on the sphere, in particular, the dynamical core of the Geophysical Fluid Dynamics (GFDL) general circulation model (for details see Boehm and Lee, 2002; Feldstein, 1994; Kim and Lee, 2001). The model circulation is driven by adding a term representing acceleration of the zonal wind by momentum flux convergence to the model zonal momentum equation.

Figure 3a shows the global steady-state vertical velocity field for a run forced with the full January momentum flux convergence field shown in Figure 2. As expected, this field agrees reasonably well with observations in the mid-latitudes, while significant differences are found in the tropics and subtropics, primarily due to the absence of the thermally driven circulation. This general feature assures the adequacy of the model setup for the investigation at hand.

Because our primary interest is on the rising motion that develops near the tropical tropopause in response to Rossby wave momentum flux convergence driven by tropical convection, we focus on model results in the tropical upper troposphere and lower stratosphere. Figure 3b shows a close-up view of the vertical velocity in this region. At the equator, rising motion is observed above about 14 km, with a maximum vertical velocity of about 0.4 mm s^{-1} at an altitude of 16.5 km. The magnitude of this rising motion is in good agreement with the estimates for this region described earlier, suggesting that tropospheric eddy momentum forcing alone is able to drive the equatorial upwelling and

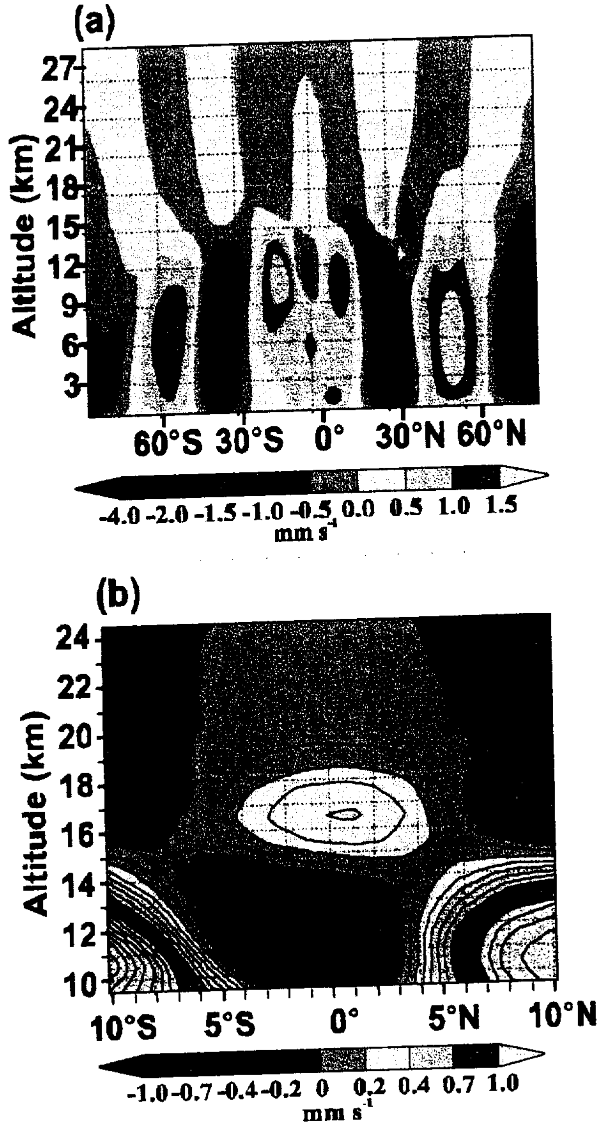

Figure 3: Vertical velocity fields for model run driven by momentum flux convergence field shown in Figure 2. (a) Global view. (b) Close-up of the tropical upper troposphere and lower stratosphere.

associated vertical moisture transport.

To investigate the relative importance of tropical and mid-latitude momentum flux convergence to the generation of tropical upwelling, runs were conducted in which the model was driven with various portions of the momentum flux convergence field shown in Figure 2. In particular, the eddy momentum flux convergence field was partitioned into tropical and extratropical regions. These runs show that in the zonal mean Rossby waves generated by tropical convection plan an equal or slightly greater role in generating tropical upwelling

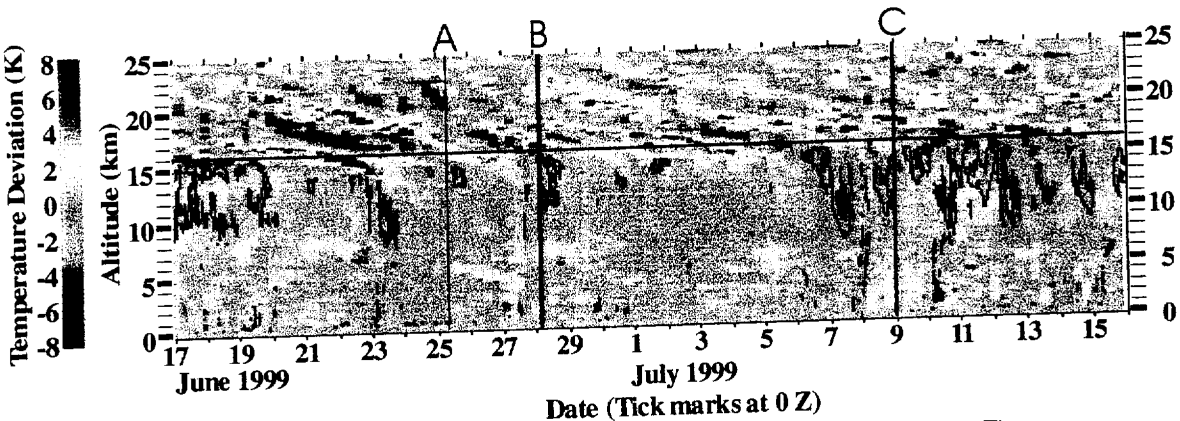

Figure 4: Results from Nauru99 showing the impact of stratospheric tropical waves on tropical cirrus. The background color plot is a time series of profiles of temperature perturbations from the Nauru99 period mean. Cloud occurrence is indicated by the superimposed black contours. See text related to Figure 6 for explanation of the horizontal and labeled vertical lines. Adapted from Boehm and Verlinde (2000).

than mid-latitude disturbances, as both mechanisms generate rising motion near the tropical tropopause on the order of 0.2 to 0.3 mm s^{-1}. However, we suspect that rising motion generated in response to Rossby waves generated by tropical convection is more effective in the vertical transport of moisture to the cold-point tropopause, since it will be more closely associated with tropical convection.

Similar model runs were also conducted for the months of April, July, and October. Although some seasonal variability was found, the magnitude of the tropical upwelling across the tropopause remains relatively constant through the course of the year.

3. LARGE-SCALE COOLING

Once moisture is present near the cold-point tropopause, large-scale rising motion and associated adiabatic cooling is required to initiate the ice crystal nucleation and growth that leads to the *in situ* formation and maintenance of tropopause cirrus on the large spatial scales that are observed. One possible source of this rising motion is the gentle ascent that is hypothesized to play a role in vertical moisture transport. Runs conducted using a one-dimensional version of the cloud resolving cirrus model developed by Lin (1997) suggest that in localized regions this rising motion may be sufficient. However, overall it appears that another source of rising motion is required to explain the observed properties and lifecycles of tropopause cirrus layers.

Stratospheric tropical waves (e.g. Kelvin and mixed Rossby-gravity waves) are an important source of temperature variability at the tropical cold-point tropopause (Tsuda et al., 1994). It has been suggested that this cooling plays a role in the life-cycle of tropopause cirrus (Jensen et al., 1996b). To test this hypothesis, Boehm and Verlinde (2000) used observations of tropical waves and cirrus occurrence from Nauru99, a field experiment conducted by the Department of Energy (DOE) Atmospheric Radiation Measurement (ARM) program on and near the island of Nauru, located less than a degree from the equator in the western Pacific Ocean.

First, a time series of temperature perturbations from radiosonde data was used to search for waves. This time series is shown by the color contours in Figure 4. It was constructed by fitting temperature data from each of the radiosondes (4 per day from 17 June 1999 through 15 July 1999) to a regulary spaced profile with 50 m spacing, combining the individual profiles into a time-altitude grid, calculating the period average at each altitude, and finally calculating the temperature perturbation at each time altitude point from the period average at that altitude. Stratospheric tropical waves with downward phase propagation were visible as coherent structures in the perturbation temperature field descending from the lower stratosphere with periods of several days and amplitudes up to 8 K. Using spectral analysis, these waves were identified as eastward propagating Kelvin waves (Holton et al., 2001). Furthermore, the properties of these waves were shown to be consistent with generation by deep tropical convection.

To look at the correlation between the Kelvin waves and tropopause cirrus, data on cloud occurrence during Nauru99 was superimposed on the temperature perturbation field. This data was obtained using a cloudmask algorithm developed by Clothiaux et al. (1998). This algorightm uses micropulse lidar data to construct a mapping of cloud occurrence. The cloudmask data is represented by black contours superimposed on the temperature data in Figure 4. Careful inspection reveals that during this period upper tropospheric thin cirrus formation occurs almost exclusively in the cold phases of the waves descending from the lower stratosphere. To further visualize this relationship, Figure 5 shows a histogram of temperature perturbations coinciding with cloud occurrences above 15 km. With few exceptions, the temperature perturbations are negative, with a peak in

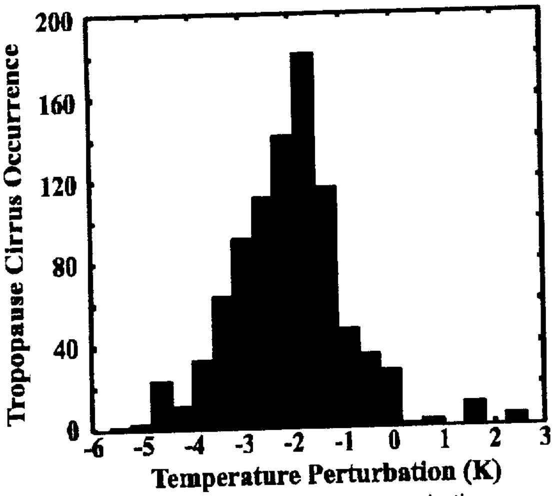

Figure 5: Histogram of temperature perturbations coinciding with cloud occurrences above 15 km during Nauru99. From Boehm and Verlinde (2000).

cloud occurrence when the temperature perturbation is about -2K, verifying the close relationship between cirrus and negative temperature perturbations near the tropopause. This result suggests that stratospheric tropical waves are an important source of cooling leading to the formation of tropopause cirrus. Furthermore, the periods (5 days and 10 days) and wavelengths (10,000 km and 20,000 km) of the waves observed during Nauru99 are consistent with typical observed life times and spatial extents of tropopause cirrus.

Global analysis data produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) has previously been used in studies of waves in the tropical upper troposphere (Randel, 1992; Fujiwara et al., 2001; Fujiwara and Takahashi, 2001; Straub and Kiladis, 2002). Here, we use ECMWF analysis data with 1°x1° spatial and 6-hour temporal resolution to examine the large-scale structure of the Kelvin waves observed during Nauru99 and also to investigate the occurrence of Kelvin waves during other periods and at other locations.

Figure 6 shows a plot of the temperature deviation from the period mean at 100 mb over the equator as a function of longitude and time for the period 14 June 1999 through 18 July 1999. This period corresponds to the period shown in Figure 4 with a couple of days added at the beginning and end. Coherent, eastward-propagating structures are visible in this temperature perturbation field. Examples of these structures are highlighted by the labeled, slanted lines. These structures correlate well with downward propagating structures in Figure 4. To show this, we have drawn in Figure 4 a horizontal line at 16 km (approximately corresponds to the 100 mb surface plotted in Figure 6) and vertical lines A, B, and C at the times when the structures indicated by lines A, B, and C in Figure 6 pass Nauru (indicated

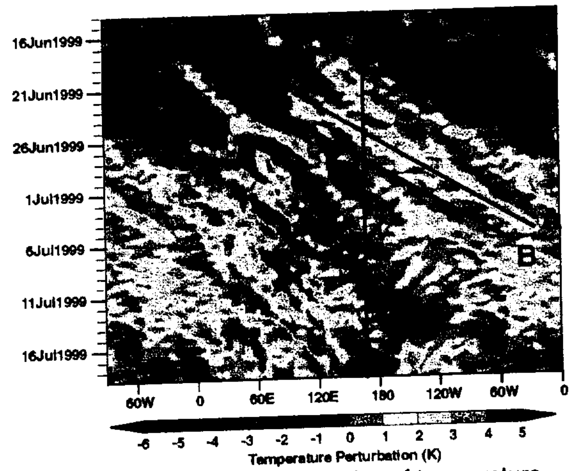

Figure 6: Longitude-time distribution of temperature perturbation from the period mean (at a given longitude) at 100 mb over the equator from ECMWF analysis data for the period 14 June 1999 through 18 July 1999. The vertical line shows the longitude of Nauru. See text for explanation of the labeled, slanted lines.

by the vertical line in Figure 6).

For example, consider the warm perturbation traced by line A in Figure 6. This perturbation passes Nauru on June 25. In Figure 4, this structure corresponds to the warm perturbation descending from the lower stratosphere from 23 km on June 19 to 16 km on June 25. Similarly, the structures traced by lines B and C are visible in Figure 4.

The zonal propagation velocity of structure A was calculated from Figure 6 to be about 23 m s^{-1}. This agrees well with the velocity calculated from the Nauru99 radiosonde data using the dispersion relationship for Kelvin waves (Holton et al., 2001). This agreement, along with the eastward propagation of the structures, strongly suggests that the structures visible in Figure 6 correspond to the Kelvin waves observed during Nauru99.

4. SUMMARY

The results presented here show that tropical convection plays a role in each of the three primary processes involved in the *in situ* formation of tropopause cirrus. First, tropical convection transports moisture from the surface into the upper troposphere. Second, tropical convection excites Rossby waves that transport zonal momentum toward the ITCZ, thereby generating rising motion near the equator. This rising motion helps transport moisture from where it is detrained from convection to the cold-point tropopause. Finally, tropical convection excites vertically propagating tropical waves (e.g. Kelvin waves) that provide one source of large-scale cooling near the cold-point tropopause, leading to tropopause cirrus formation.

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